DISTRIBUTION OF CRUSTAL MAGNETIC FIELDS ON MARS: SHOCK EFFECTS OF BASIN-FORMING IMPACTS.

L. L. Hood, N. C. Richmond, *Lunar and Planetary Laboratory, University of Arizona, Tucson AZ USA, (lon@lpl.arizona.edu)*, E. Pierazzo, *Planetary Science Institute, Tucson, Arizona USA*, P. Rochette, *University of Aix-Marseille 3, Aix en Provence, FRANCE*.

Introduction. Crustal magnetic fields on Mars are inhomogeneously distributed (Figure 1a) with the strongest fields occurring over the southern highlands in a longitude sector between approximately 130°E and 240°E (1). Here we investigate whether radially extended shock demagnetization associated with the Hellas and Argyre basin impacts (in addition to thermal demagnetization associated with the northern resurfacing event and the formation of the Tharsis volcanic complex) may largely explain this distribution (2).

Approximate Models for Shock Pressure Decay. A simple approximation of shock pressure decay with radial distance can be obtained using an empirical determination of particle velocity decay and the Hugoniot equations, (e.g. ref. 3, p. 66): $P(r) = \rho_{\circ t}[C_t + S_t u_\circ (r_\circ/r)^{1.87}] u_\circ (r_\circ/r)^{1.87}$, where $\rho_{\circ t}$ is the unshocked target mass density, $u_\circ$ is the initial particle velocity at distance $r_\circ$ from the impact point, and C_t, S_t are empirically determined shock parameters (3). Using the planar impact approximation, Ahrens and O'Keefe (4) estimated peak particle velocities of 3.75 and 7.5 km/s for a gabbroic anorthosite projectile impacting on a gabbroic anorthosite target at 7.5 and 15 km/s respectively. Substituting x for $(r_\circ/r)^{1.87}$, this equation becomes a simple quadratic equation, that can then be solved for the normalized radial distance at which the shock pressure is equal to a given P_1. For example, for $P_1 = 2$ GPa, this gives $r/r_\circ = 8.8$ for a 7.5 km/s impact, and $r/r_\circ = 12.7$ for a 15 km/s impact. Here, we have used $\rho_{\circ t} = 3.965$ kg/m^3, $C_t = 7.71$ km/s and $S_t = 1.05$ for a gabbroic anorthosite (high pressure phase) composition (Table 4.2 of ref. 3).

To convert these relative distance estimates to absolute distances from the centers of the Hellas and Argyre basins, it is necessary to estimate the radii of impactors that could have produced these basins. Assuming that the transient crater diameter coincides with the inner topographic boundaries (based on MOLA data; Fig. 1b) of the Hellas and Argyre basins, $\sim$ 1400 and $\sim$ 1000 km, respectively, the π-scaling law (5) yields a projectile diameter between 685 and 463 km for Hellas, and between 445 and 301 km for Argyre, for impact velocitites between 7.5 and 15 km/s. (For these calculations we used the impact scaling code developed by H.J. Melosh, available at www.lpl.arizona.edu/tekton/crater.html; other input parameters used are: 2900 kg/m^3 for impactor and target density (appropriate for a low pressure phase gabbroic anorthosite composition); 3.72 m/s^2 for the acceleration of gravity; 45^o for most probable impact angle.) Thus, the 2 GPa shock pressure is reached between 2940 and 3000 km from the center for a 7.5 to 15 km/s Hellas–forming impactor, and between 1911 and 1949 km away for an Argyre-forming impactor.

In Figure 1b, the inner contours represent the mean distance from the structure's center at which the peak shock pressure should have reached approximately 6 GPa plotted on the MOLA Mercator projection. The 4 GPa and 2 GPa contour lines are similarly shown in Figure 1b. To simplify comparisons with the magnetic field distribution, the pressure contours are also replotted in Figure 1a.

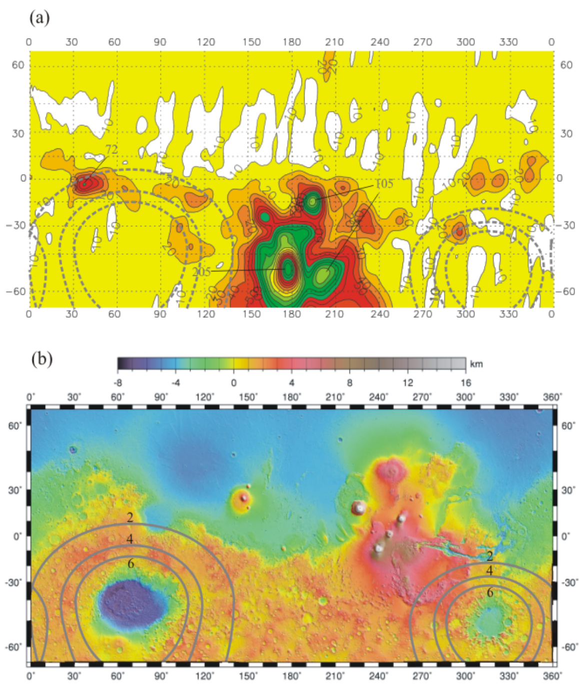

Figure 1:

Remanence Carriers and Shock Demagnetization. Based in part on studies of shergotite, nakhlite, and chassignite (SNC) meteorites, several candidate crustal remanence carriers on Mars have been proposed. These include iron oxides such as titanomagnetite and hematite (6,7,8) and iron sulfides, i.e., monoclinic pyrrhotite (9). An important problem with titanomagnetite as the main magnetic carrier for martian crustal magnetic anomalies is its low Curie temperature (150°C), which would inhibit magnetization at depths of more than $\sim$ 10 km during early martian history. In contrast, studies of martian magnetic spectra as well as the absence of demagnetization signatures for smaller craters suggest depths for martian magnetic anomaly sources of up to 30 - 50 km (10,11).

The efficiency of shock demagnetization during the Hellas and Argyre impacts depends partly on the identity of the primary magnetic remanence carrier. If iron oxide remanence carriers are assumed, then results of shock demagnetization experiments on terrestrial basalt samples should be approximately applicable to the martian case. For example, the experiments of Pohl et al. (12) show that shock demagnetization of titanomagnetite grains is effective mainly for those grains with low coercivities (stability against alternating field demagnetization). Shock stresses of 0.25 GPa are sufficient to demagnetize such grains with coercivities of 5 to 10 mT while stresses of 0.8 GPa are sufficient to demagnetize grains with coercivities of 15 to 20 mT. However, the remanent coercivities of SNC meteorites containing titanomagnetite remanence carriers range from 55 to 63 mT (e.g., Table 1 of Rochette et al. (9)). It is therefore likely that much larger shock stresses would be needed to demagnetize these samples if titanomagnetite is the main remanence carrier. Experiments by Hargraves and Perkins (13) on terrestrial samples containing homogeneous magnetite as the main magnetic phase indicate that shock pressures $>$ 5 GPa are necessary to significantly reduce the remanence intensity. The latter authors also report no effect of impact shock on α hematite in samples of Moenkopi red beds collected around Meteor Crater, Arizona. To our knowledge, no experimental data are available on shock demagnetization of rocks containing hematite-ilmenite lamellae (8) or magnetite-ilmenite (14). If, on the other hand, high-coercivity monoclinic pyrrhotite is assumed to be the main remanence carrier, then nearly complete demagnetization is expected at pressures near 2.7 GPa (15). A pyrrhotite-bearing sample shows a decrease of 58% of its initial remanence after compression to 1 GPa at room temperature. Hence, extensive ($\sim$ 90%) demagnetization is expected at pressures of order 2 GPA and significant ($\sim$ 50%) partial demagnetization occurs at pressures of 1 GPa or less.

Discussion and Conclusions. Recently, a partial correlation between magnetic anomalies and valley networks, which are believed to have been caused by surface water erosion, has been reported (16). This characteristic suggests that magnetization in the deep martian crust may have been stimulated in part by hydrothermal circulation that also led to erosive formation of the valley networks. The distribution of valley networks (17) is characterized by a broad arc that roughly follows the series of strong anomalies in the equatorial region (see Figure 10 of ref. 16). Therefore, the presence of these weaker anomalies in the equatorial zone outside of the 130°E to 240°E sector may be related to hydrothermal alteration processes in the deep crust. However, valley networks are widely distributed in longitude and are equally numerous in both hemispheres. Therefore, this mechanism alone can not explain the concentration of strong anomalies in the 130°E to 240°E sector.

Results of the approximate peak shock pressure calculations shown in Figure 1 combined with available experimental data on magnetic remanence versus peak shock pressure for likely martian remanence carriers suggest that the concentration of strong anomalies in the 130°E to 240°E sector can potentially be attributed to impact shock demagnetization. For example, if pyrrhotite is assumed to be the major remanence carrier in the martian crust, then extensive ($\sim$ 90%) demagnetization may be expected within 3-4 basin radii. The remaining strong magnetization centered on 180° longitude may therefore be a surviving remnant of the early Noachian crust that escaped both impact demagnetization and thermal alteration processes.

References. (1) Acuña, M. et al., *Science*, *284*, 790, 1999; Connerney et al., *Science*, *284*, 794, 1999; Acuña et al., *J. Geophys. Res.*, *106*, 23403, 2001; Hood, L. L., N. C. Richmond, E. Pierazzo, and P. Rochette, *Geophys. Res. Lett.*, in press, 2003; (3) Melosh, H. J., *Impact Cratering: A Geologic Process*, 245 pp., Oxforc Univ. Press, New York, 1989; (4) Ahrens, T. J., and J. D. O'Keefe, in *Impact and Explosion Cratering*, ed. by D. J. Roddy et al., p. 639, Pergamon, New York, 1977; (5) Schmidt, R. M. and K. Housen, Some recent advances in the scaling of impact and explosion cratering, *Int. J. Impact Eng.*, *5*, 543-560, 1987; (6) Hargraves, R. et al., *J. Geophys. Res.*, *82*, 4547, 1977; (7) Kletetschka, G. et al., *Meteorit. Planet. Sci.*, *35*, 895, 2000; (8) Robinson et al., *Nature*, *418*, 517, 2002; (9) Rochette, P. et al., *Earth Planet. Sci. Lett.* *190*, 1, 2001; (10) Voorhies, C. V. et al., *J. Geophys. Res.*, *107(E6)*, p. 1-1 to 1-10, 2001JE001534, 2002; (11) Nimmo, F., and M. S. Gilmore, *J. Geophys. Res.*, *106*, 12315, 2001; (12) Pohl, J. et al., *J. Geophys. Res.*, *41*, 23, 1975; (13) Hargraves, R. B. and W. Perkins, *J. Geophys. Res.*, *74*, 2576, 1969; (14) Nimmo, F., *Geology*, *28*, 391, 2000; (15) Rochette, P. et al., *Lunar Planet. Sci. XXXIII*, Abstract 1199, LPI, Houston, 2002; (16) Harrison, K. P. and R. E. Grimm, Controls on martian hydrothermal systems: Application to valley network and magnetic anomaly formation, *J. Geophys. Res.*, *107(E5)*, doi:10.1029/2001JE001616, 2002; (17) Kieffer, H. H., in *Third International Colloquium on Mars, LPI Contrib. 441*, p. 133, LPI, Houston, 1981.